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Tropical cooling and the onset of North American glaciation

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Abstract

We offer a test of the idea that gradual cooling in the eastern tropical Pacific led to cooling of North America and the initiation of glaciation ~3 Myr ago. Using modern climate data we estimate how warming of the eastern tropical Pacific affects North American temperature and ice-ablation. Assuming that the modern relationship holds over the past millions of years, an eastern tropical Pacific warmer by ~4° between 3–5 Ma would increase ablation in northern North America by approximately two meters per year. By comparison, a similar estimate of the ablation response to variations in Earth’s obliquity gives less than half the magnitude of the tropically-induced change. Considering that variations in Earth’s obliquity appear sufficient to initiate glaciations between ~1–3 Ma, we infer that the warmer eastern equatorial Pacific prior to 3 Ma suffices to preclude glaciation.

1 Introduction

Milankovitch (1941) hypothesized that when the Northern Hemisphere is relatively distant from and less pointed toward the sun during summer, ablation is weak and Northern Hemisphere icesheets grow – and that under less favorable orbital conditions, icesheets decay. Elaborated upon in the following decades, this hypothesis nonetheless remains the predominant view regarding the control of glaciation (e.g. Imbrie et al., 1992). Such glacial cycles are, however, the exception when considering the whole of Earth’s history (e.g. Zachos, 2001). Indeed, for glacial cycles to occur over the past few million years the climate would seem to be precariously poised between glaciated and unglaciated states, particularly if icesheets are to wax and wane in response to the relatively weak redistribution of insolation induced by orbital variations – no more than some tens of W/m² and usually much less. Milankovitch’s hypothesis suggests that glaciation would be precluded if summer ablation were greater by an amount equivalent to that associated with orbital variations. Here we consider whether warmer tropi-

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cal sea-surface temperature can account for the apparent absence of glacial cycles in North America prior to 3 Ma.

There is a growing interest in whether the tropical climate could change high-latitude conditions (e.g. Cane, 1998). In particular, a change in tropical SSTs may be capable of altering climates at high latitudes through atmospheric teleconnections (Cane and Molnar, 2001; Philander and Fedorov, 2003; Molnar and Cane, 2002). Recent estimates of eastern Pacific SSTs using $\delta^{18}\text{O}$ (e.g. Ravelo et al., 2004; Haywood et al., 2005), Mg/Ca (e.g. Wara et al., 2005; Ravelo et al., 2006), and alkenones (e.g. Lawrence et al., 2006) show that temperatures were $\sim 4^\circ\text{C}$ warmer between ~ 3 to 5 Ma than is typical today (but see Rickaby and Halloran, 2005 for a different view), and $\delta^{18}\text{O}$ and Mg/Ca measurements also indicate that western tropical Pacific SSTs were cooler by $\sim 1^\circ\text{C}$ (Wara et al., 2005; Ravelo et al., 2006). The implied flattening of the tropical Pacific SST gradient between ~ 3 to 5 Ma is thus similar to, if more complete than, the SST relaxation during typical, modern El Niño events. Thus, the idea has emerged that in pre-iceage times the eastern tropical Pacific Ocean was in a permanent El Niño state, with a warm SST extending across the Pacific, and with extratropical climates mimicking those associated with El Niño teleconnections (e.g. Fedorov et al., 2006; Molnar and Cane, 2002; Philander and Fedorov, 2003). The warming of much of Canada and Alaska during El Niño events suggests that a warm eastern equatorial Pacific Ocean may explain the absence of continental ice sheets in the period preceding ~ 3 Ma (e.g. Barreiro et al., 2006).

Haywood et al. (2007) challenged this scenario of tropical inhibition of glaciation. They explored the sensitivity of a coupled GCM to changes in tropical temperatures and concluded that permanent El Niño would have a small influence on North American glaciation on the basis of small changes in the global annual mean temperature. More locally, however, a permanent El Niño in their model appears to warm the high-latitudes in North America by as much 5°C in the winter and 2°C in the summer (Haywood et al., 2007, Fig. 1b) – consistent with other studies (Barreiro et al., 2006). Investigation of the local North American temperature response, as well as the structure of the full

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seasonal cycle, seems necessary prior to drawing conclusions regarding El Niño's influence on glaciation. Furthermore, as simple as a warmer eastern tropical Pacific may seem for accounting for the absence of icesheets before ~3 Ma, it needs to be reconciled with a fundamental component of Milankovitch's hypothesis — that summer is the most important season. El Niño teleconnections are clearly documented to cause winter warming in North America (e.g. Ropelewski and Halpert, 1986; Trenberth et al., 1998), but to increase ablation of ice, summer – not winter – must warm. This conundrum motivates us to explore the influence of El Niño on the full seasonal cycle of temperature in North America and then to extrapolate this relationship to past climates.

Our focus is on eastern tropical Pacific temperatures but, of course, other features of the climate could also contribute to, or themselves explain, the absence of glaciation in North America during the early Pliocene. Suggestions include higher concentrations of greenhouse gases (e.g. Haywood et al., 2005) and greater ocean transport of heat (e.g. Rind and Chandler, 1991). But the strong evidence that the eastern tropical Pacific was warmer during the early Pleistocene prompts our consideration of whether this change alone could account for the absence of North American glaciation. The calculation is in the spirit of what might be done on the back of envelope – with the aim of estimating whether eastern equatorial Pacific cooling could lower summer ablation sufficiently to permit initiation of glaciation.

2 High-latitude temperature responses to El Niño

We begin by making an empirical estimate of the high-latitude temperature response to El Niño events using instrumental data. The analysis relies mostly upon the Global Historical Climatology Network (GHCN) (Peterson and Vose, 1997), which provides daily resolution instrumental temperatures and has recently been updated. As the GHCN records give only daily maximum and minimum temperature, we use the average of these two to estimate daily average temperature. Note that El Niño events are associated with small variations in precipitation and snow accumulation at high-latitudes, so

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that discussion is focused on temperature. The extra-tropical temperature response is analyzed in three steps:

Identification: El Niño events are identified by sea-surface-temperature anomalies (Kaplan et al., 1998) averaged over the NINO3.4 region – that is, between 120° W and 170° W and within 5° latitude of the equator. After smoothing temperature anomalies with a five-month running mean, we selected the ten largest temperature excursions. These “El Niño intervals” are defined to begin in January of 1877, 1888, 1902, 1905, 1940, 1972, 1982, 1986, 1991, and 1997 and to extend for two years.

Extra-tropical response: We average as many of the ten El Niño intervals as are present in each GHCN temperature record into a single, composite El Niño interval. Only records from the northern extra-tropics that span at least one El Niño event are included, giving 363 records with, on average, 62 years of daily observations.

Time/space description: A singular value decomposition of the 363 composite El Niño intervals characterizes the temporal and spatial extra-tropical temperature response to an El Niño event. To isolate low-frequency fluctuations, we first smooth each composite using a 90 day tapered running average and truncate the first and last 45 days of the record. The first three principal component/empirical orthogonal function pairs (PC/EOF) are readily interpreted (see Fig. 1). The leading PC/EOF pair, accounting for 41% of the temperature variance, indicates warming between December and March and is concentrated in the interior of North America at mid-latitudes. Winter temperature can breach the freezing point at mid-latitudes, and warming is expected to increase ablation.

The second PC/EOF pair, which explains 17% of the temperature variance, indicates warmer temperatures throughout northern North America during the second year of the El Niño interval. This sustained high-latitude warming of North America appears the most pertinent with regard to glaciation. We checked this relationship further by comparing the NINO3.4 index with northern North American surface air temperatures from the Climate Research Unit’s monthly resolved temperature compilation (Jones and Moberg, 2003). (To be specific, we define the region between 130° W and 70° W

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and north of 50° N as the high-latitudes or northern North America.) These data reach back to 1885, earlier than most GHCN records. For the NINO3.4 index we took the annual average for the period from June to May (the warmest El Niño months), and for the surface air temperatures we used the average between January and December (consistent with the warm phase of the second PC). After detrending both time-series, the cross-correlation between the surface air temperature and NINO3.4 between 1885 and 2005 is 0.49 (Fig. 2). Monte Carlo experiments, accounting for the autocorrelation in both time-series, indicate this correlation is significant at the 99% confidence level. Regression between the NINO3.4 index and high-latitude temperature gives a slope of 0.51, but this is biased because of the presence of noise in both time-series. A simple correction is to divide the regression slope by the cross-correlation (see e.g. Frost and Thompson, 2000), giving a bias corrected slope of 1.

The third PC/EOF, which accounts for 12% of the variance, is the final useful descriptor of temperature variability – subsequent PC/EOF pairs explain less than 6% of the variance. The third PC/EOF most closely resembles the time-evolution of the NINO3.4 temperature during an El Niño event – excepting the January to February negative excursion. Measures of teleconnections with El Niño that rely upon a linear fit with NINO3.4 seem to emphasize this mode, giving the incorrect impression that El Niño warming in North America is confined to the margins (see e.g. Halpert and Ropelewski, 1992). In actuality, PC/EOF pairs one and two show not only that warming is greatest in the interior of North America, but also that some warming occurs during the spring, summer, and fall – suggesting that the eastern tropical Pacific could influence glaciation. A similar analysis performed using only North American temperature records yields these same conclusions.

3 Temperature and ablation

Surface temperature reconstructions in the eastern Equatorial Pacific indicate a long-term cooling trend of approximately four degrees Celsius since four million years ago

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(e.g. Ravelo et al., 2004; Wara et al., 2005; Lawrence et al., 2006). The simplest approach to estimating the extra-tropical response is to assume that in the past the eastern equatorial Pacific was uniformly warmer throughout the year, and that the annual average warming observed for NINO3.4 is proportional to a constant temperature change in the extra-tropical Northern Hemisphere. We estimate the proportionality as the ratio of the annual average (December to November) land temperature anomaly to the annual average (June to May) NINO3.4 temperature during the ten largest El Niño events. The NINO3.4 and northern North American temperature anomalies both average $\sim 1^\circ\text{C}$, giving a ratio of ~ 1 (Fig. 3), as might be expected from EOFs one and two and from the regression between the NINO3.4 index and average temperature in northern North America.

High-latitude temperature anomalies, of themselves, do not indicate changes in ablation simply because warmer temperatures that are still sub-freezing do not much affect ablation. To quantify the influence of daily temperature on ablation we use positive degree days (PDDs),

$$\text{PDD} = \sum_{d=1}^{365} \delta_d T_d. \quad (1)$$

T_d is the daily temperature and δ_d is zero if $T_d < 0^\circ\text{C}$ and one otherwise. The composite El Niño intervals, however, come from averages over different years, and the distribution of temperature around this average needs to be accounted for (see Braithwaite, 1984). For example, if the average temperature on the first of January hovers just below freezing, the expected PDDs are nonetheless positive because some realizations of January first temperature will inevitably exceed freezing. We assume a normal distribution with a standard deviation calculated from, for example, all January first temperatures in a record. The expected number of PDDs becomes,

$$\langle \text{PDD} \rangle = \sum_{d=1}^{365} \int_0^\infty x \gamma(\bar{T}_d, \sigma_d, x) dx. \quad (2)$$

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y is a normal distribution of temperature with mean \bar{T} and standard deviation σ on day d , and x is temperature in Celsius integrated from the freezing point upward. All PDDs reported in this paper are the expected values, but with the brackets left off for simplicity. For stations where seasonal temperatures hover near freezing, PDDs calculated using Eq. (2) exceed those using Eq. (1) by as much as 50%.

PDDs tend to be smallest at high-latitudes and in the interior of continents. A one degree annual average rise in the NINO3.4 index – corresponding to a moderate El Niño event – is associated with, on average, a 90°C-day increase in northern North America (Fig. 3d). Note that peak differences in PDDs lie south of peak differences in temperature because more days breach the freezing point to the south. PDDs increase super-linearly with temperature because both the magnitude increases and the number of days with positive temperatures increases. Thus, assuming that extra-tropical temperatures continue to scale linearly with the NINO3.4 index, a 2°C warming in the NINO3.4 region is associated with a 300°C-day increase; a 4°C tropical warming gives a 630°C-day increase, but this pushes the limits of plausibility in extrapolating El Niño events to other climates. In Eurasia, during El Niño events, PDDs decrease slightly in the interior and increase slightly near the margins.

Common scaling factors relating ablation rates (in terms of water equivalent thickness) to PDDs range from 3 mm/(°C-day) for snow-covered surfaces to 8 mm/(°C-day) for bare ice (Braithwaite and Zhang, 2000; LeFebvre et al., 2002). The higher rate of ice ablation owes to a lower albedo. Thus a 300 °C-(day) increase in northern North America corresponds to an ablation rate higher by one to two meters per year than today. Considering that for the modern climate northern North America receives, on average, 0.4 meters per year of precipitation (according to GHCN records, Peterson and Vose, 1997), such a large increase in ablation would presumably inhibit glaciation.

So far we have assumed that the response to eastern tropical Pacific warming is constant through the seasons, as a simplest possible case, but a seasonally varying response is perhaps more realistic. Furthermore, observation of past temperatures are spaced at centennial or lower resolution, and ENSO variability, if it is present, cannot

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be directly resolved. Thus we cannot distinguish whether the long-term decline in eastern equatorial temperatures owes to a shift in mean temperature or to decreases in the frequency or amplitude of El Niño events (e.g. Haywood et al., 2007). To explore the implications of seasonal temperature variability – as opposed to a constant response across the seasons – we perturbed each extra-tropical temperature record by a seasonally varying temperature anomaly. Specifically, we add to the temperature on each day of a record the temperature anomaly on the corresponding day in the composite El Niño interval, taken between December to November. The resulting change in PDDs is actually somewhat larger for northern North America than that obtained from the annual average scaling described earlier, indicating that an increase in ablation is a robust implication of the data.

4 Comparison with insolation

The magnitude of the tropically-induced changes in PDDs can be placed in context by comparison with the effects of Earth's orbital variations. If changes in Earth's orbital configuration, and the resulting shifts in insolation, forced North America between unglaciated and glaciated states, then an estimate of the magnitude of the difference in ablation between glacial and interglacial periods indicates the forcing required to initiate a glaciation. Two factors complicate simple application of this logic. First, glacial cycles might occur independent of orbital variation, perhaps only being paced by changes in insolation (e.g. Tziperman et al., 2006). Such self-sustained oscillation seem most relevant for the ~100 Ky late Pleistocene glacial cycles, but the early Pleistocene 40 Ky glacial cycles appear to respond directly to radiative forcing, with the timing and amplitude of changes in glaciation closely following the integrated summer energy (Huybers, 2006). As the integrated summer energy is almost wholly a function of obliquity, we use the difference in PDDs between low and high obliquity to gauge the forcing required to preclude glaciation.

The second complication is that feedbacks – for example, involving ice-albedo,

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icesheet instabilities, greenhouse gases, or ocean and atmospheric circulation shifts – could amplify the glacial response to insolation variability. Furthermore, Liu and Herbert (2004) found that eastern tropical Pacific temperatures covary with obliquity and the extent of glaciation, confounding separation of orbital and tropical Pacific influences on glaciation. Insofar as climate feedbacks contribute only after ice is growing, however, their effects are irrelevant with respect to the onset of glaciation.

Earth's orbital variations are well-known over the last 5 Ma (e.g. Berger and Loutre, 1992), and these could be specified within a general circulation model (e.g. Jackson and Broccoli, 2003) or, more simply, an energy balance model (e.g. Huybers and Tziperman, submitted) to determine the sensitivity of PDDs. Our experiments with energy balance models, however, indicate that the relationship between solar insolation and temperature can be winnowed down to three parameters – a lag, L ; sensitivity, S ; and constant, B – and we opt for a very simple representation,

$$T_i(d + L) = S_i I(d, \phi_i) + B_i, \quad (3)$$

In words, the lagged daily average temperature, $T(d+L)$, at station i is related to daily average insolation, $I(d, \phi)$, by a sensitivity, S , and a constant temperature, B . ϕ_i is the latitude at which temperature is recorded. The average cross-correlation obtained between Eq. (3) and the average seasonal cycle at each site is 0.99 (Fig. 4a) – as high as could be obtained through fitting an energy-balance-model. Eastern continental interiors show the greatest sensitivity and least lag with insolation, presumably because these regions are the most isolated from the buffering effects of the oceans.

We expect the relationship between temperature and insolation estimated using modern observations to hold under different orbital configurations, at least to first order during similarly unglaciated times because the variations in insolation are small relative to the mean and are concentrated at the seasonal cycle. For example, shifting obliquity from 22.2° to 24.2° causes an increase in the amplitude (measured from maximum to minimum intensity) of the annual cycle at 65° N of 40 W/m^2 and only a 6 W/m^2 change in the annual mean insolation. In the case of precession, shifting perihelion from winter

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to summer solstice causes an increase in the amplitude of the annual cycle by 54 W/m^2 and the annual mean insolation is invariant.

Assuming that the relationship between temperature and insolation (Eq. 3) estimated from modern data holds under different configurations of Earth's orbit, we calculate the temperature at each site for the seasonal cycles when obliquity is 22.2° and 24.2° . PDDs are calculated from temperature using Eq. (2). The specified two-degree increase in obliquity is a typical amplitude for an obliquity cycle and, according to our calculations, causes an increase of 170°C-days in northern North America. Jackson and Broccoli (2003) obtain a similar result from their GCM calculation. The difference in PDDs at each site depends on both the temperature sensitivity and the difference in insolation. Thus the PDD response to an increase in obliquity is largest in continental interiors, away from the buffering effects of the ocean, and at high latitudes where obliquity-induced difference in insolation are greatest.

5 Conclusions

Finally, we consider the combined influence of tropical temperature and orbital variations. As our interest is in discerning if the long-term eastern tropical Pacific temperature changes could have initiated glaciation, we fit a third order polynomial to the 5 Myr temperature record of Lawrence et al. (2006). This fit indicates an overall cooling of 4°C , and which we take as a proxy for a decreasing NINO3.4 index. The anomalies in the NINO3.4 proxy are used to calculate extra-tropical temperature changes, as described in Sect. 3. The orbitally-forced temperature changes over the last 5 Ma are calculated as described in Sect. 4, in this case including changes in obliquity, precession, and eccentricity (Berger and Loutre, 1992). Fig. 5 shows the variations in PDDs from the combined tropical- and insolation-forced temperature response, assuming that the modern standard deviations in temperature also hold for the past.

The inferred decrease in PDDs during the past 5 Myr is $\sim 600^\circ\text{C-days}$, much larger than the $\sim 170^\circ\text{C-days}$ associated with variations in obliquity (Fig. 5). The steepest

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tropically induced change in PDDs occurs between 3.5 Ma and 2.5 Ma, coinciding with a rapid increase in the $\delta^{18}\text{O}$ of benthic foraminifera, and which Mudelsee and Raymo (2005) associated with the onset of Northern Hemisphere glaciation. Although the modern estimate of the ratio between tropical and high-latitude warming need not hold under much warmer tropical conditions, the tropical forcing appears sufficiently strong that even were its influence decreased by a factor of two, it would still be larger than the obliquity-induced changes in PDDs.

According to Milankovitch's hypothesis, summer ablation must dip below a threshold value for glaciation to initiate. Our calculations indicate that the decline in eastern tropical Pacific temperatures since ~5 Ma would decrease ablation in northern North America by more than twice as much as orbitally-induced changes, and thus could readily bring the climate close enough to glaciation such that, during a favorable orbital configuration, an icesheet would form. We thus conclude that tropical cooling is sufficient to switch North America from an unglaciated state to one with intermittent glaciations.

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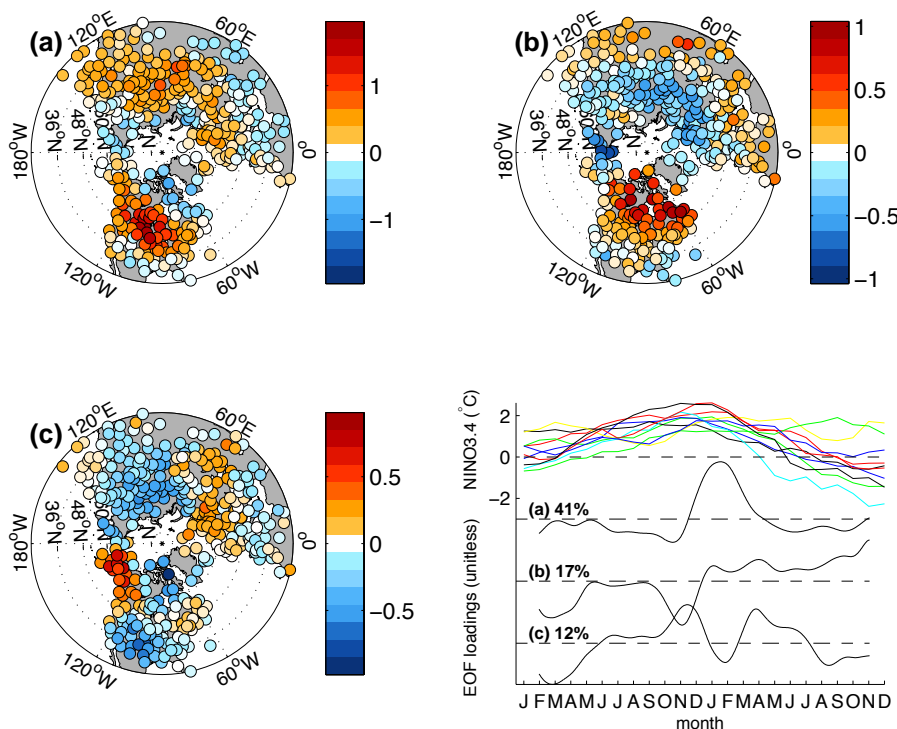


Fig. 1. Singular value decomposition of surface air temperature variability during El Niño events. **(a–c)** The spatial patterns (empirical orthogonal functions) associated with modes one through three. Shading indicates the temperature change in degrees Celsius from a one-standard-deviation increase in the corresponding principal component. **(d)** (top) The NINO3.4 temperature index during a two-year interval for the ten largest El Niño events since 1880. (below) The three leading principal components, where the dashed lines indicate the respective zero-values and the percentage of variance each principal component describes is listed.

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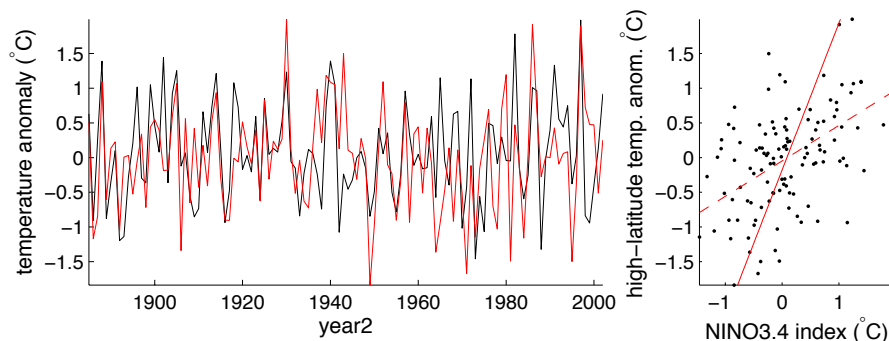


Fig. 2. (Left) Tropical Pacific (black) and high-latitude continental (red) temperature anomalies. Tropical temperatures from the NINO3.4 region are averaged between June to May and high-latitude temperatures from continental stations between 160°W and 55°W and above 50°N are averaged between December to November. Both time-series are plotted against the year associated with December. (Right) The cross-correlation between the NINO3.4 index and high-latitude temperature anomalies is 0.49. A least-squares linear fit to the data yields a slope of 0.5 (dashed line) but this is biased (see the text) and is corrected to ~ 1 (solid line).

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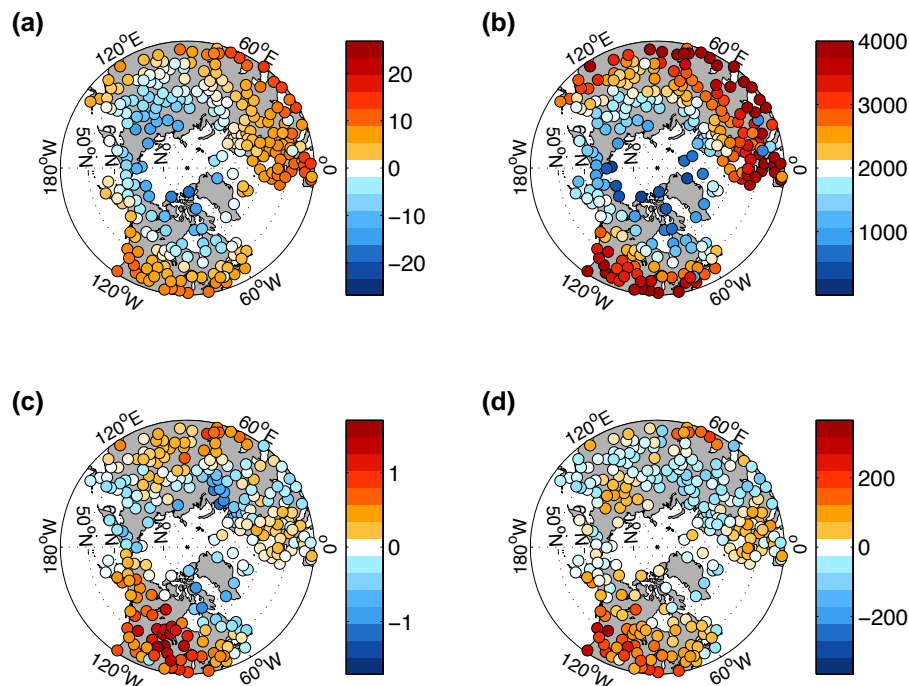


Fig. 3. (a) Northern Hemisphere temperature in $^{\circ}\text{C}$, (b) the number of positive-degree-days in $^{\circ}\text{C}$ days, and (c, d) their respective anomalies during El Niño events. Temperature and PDD anomalies are the December to November average of the ten largest El Niños intervals since 1880, or fewer, if the record begins after 1880.

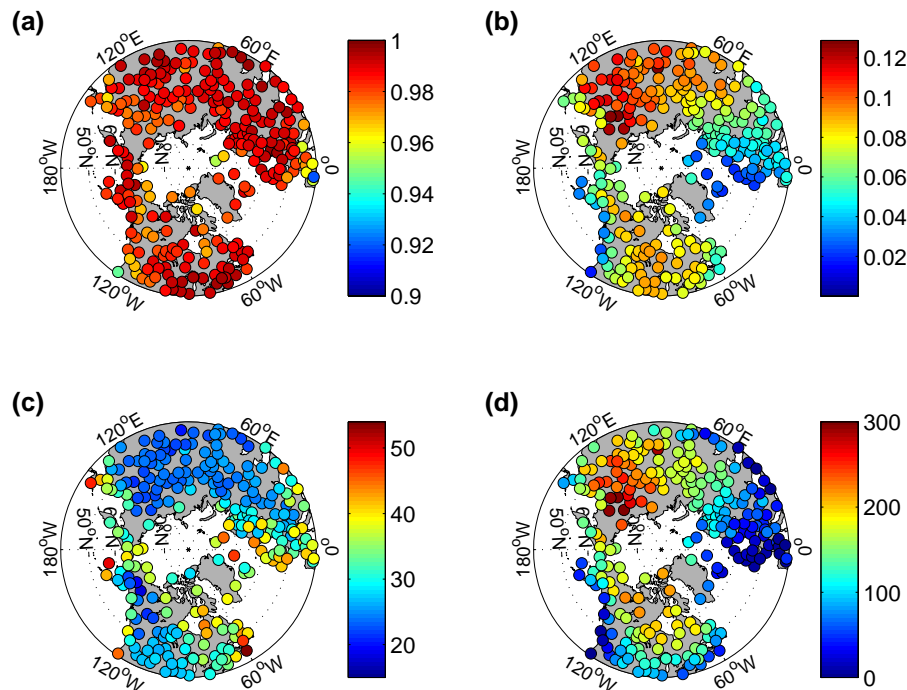


Fig. 4. The temperature and positive degree day (PDD) response to changes in insolation. **(a)** The squared-cross-correlation between lagged insolation and the average seasonal cycle in temperature. **(b)** The sensitivity of temperature to insolation, in units of $^{\circ}\text{C}/(\text{W}/\text{m}^2)$. **(c)** The number of days temperature lags incoming insolation at the top of the atmosphere. **(d)** The calculated difference in PDDs resulting from changing the the tilt of Earth's spin axis, or its obliquity, from 22.2° to 24.2° , an amount similar to the amplitude of an average obliquity cycle.

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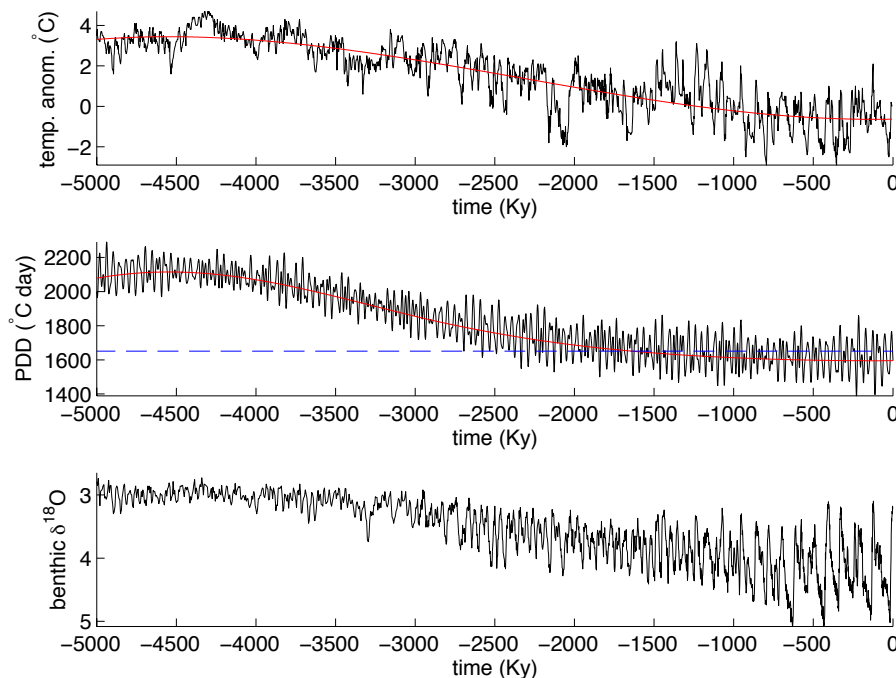


Fig. 5. Glacial implications of tropical cooling. (top) Alkenone derived SST anomalies for the eastern equatorial Pacific (from Lawrence et al., 2006), normalized so that the most recent estimate is 0°C. Red line indicates the long-term trend, fit using a third-order polynomial. (middle) Calculated average positive degree days for northern North America (between 70° W and 130° W and above 50° N) obtained by applying temperature perturbations expected to result from tropical anomalies (red line) and also including orbital variations (black line). For reference, the lowest number of PDDs obtained by 2.7 Ma is indicated (dashed line). (bottom) A combination of many benthic $\delta^{18}\text{O}$ records (Lisiecki and Raymo, 2006), where more positive values indicate greater ice-volume. Note the y-axis is reversed.

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